Gálvez-Davison Index calculation

The GDI is a stability index that performs best on the tropics and subtropics. It works best in trade wind regimes, and regions downwind such as eastern continents in tropical and subtropical latitudes. The GDI considers only moisture and temperatures at four levels. It can thus be applied into different databases such as analyses, model output, satellite-derived fields and soundings, for example.

Input data

The GDI requires temperature and mixing ratio data at four levels: 950, 850, 700 and 500 hPa. When plotting the GDI in grids, an optional correction is available to improve visualization over high mountain ranges, where high fictitious GDI values occur. If you want to apply this correction, you need surface pressure data.

General philosophy

The index considers working with three layers as indicated in Figure 1. The data is then organized into these three layers (for details just follow the equations presented later in this document). Layer A considers 950 hPa data representing near-ground conditions. Layer B considers an average of 850 and 700hPa data representing the low-mid troposphere, and layer C considers 500 hPa data representing the mid-level troposphere. You can find these layers in Figure 1, which also shows the conceptual model of convective regimes in tropical/subtropical trade wind regions and their signature on the field of equivalent potential temperature.

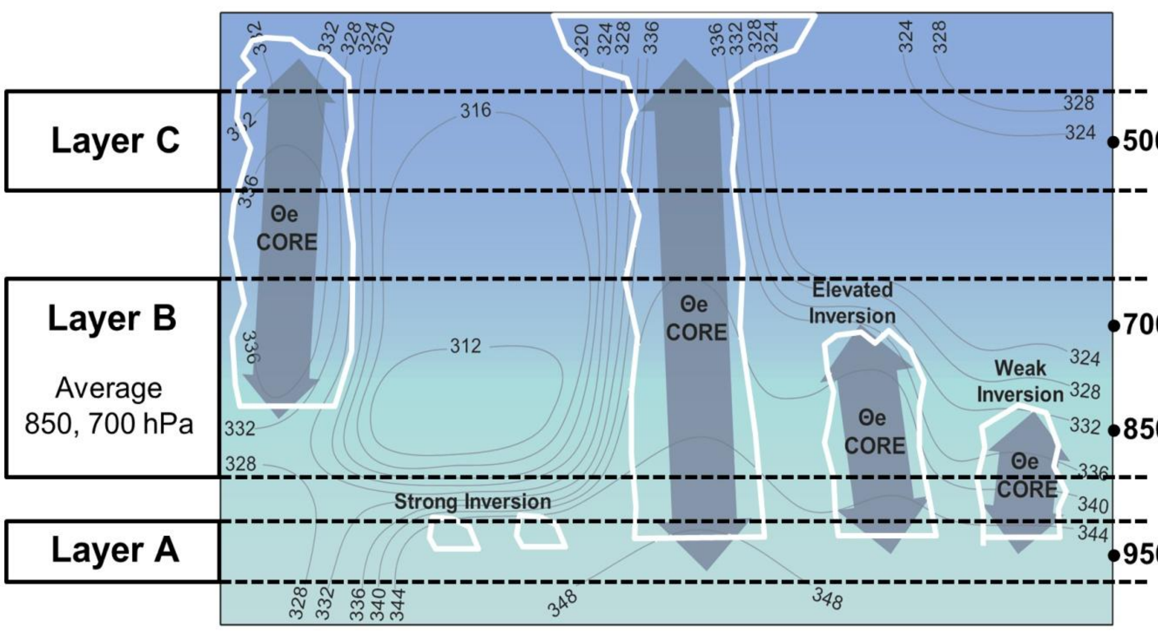

Figure 1. Conceptual model of dominant tropical convection regimes (white contours for clouds) and their signature in the θ_e field (light gray contours, data in K). The three layers used for GDI calculations and corresponding pressure levels are also included.

Calculation Algorithm

The first step is to compute potential temperatures [in K] and mixing ratios [in kg/kg] for each layer via equations 1.1 through 1.6. Input temperatures are always in [K]:

$$\theta_A = \theta_{950} = T_{950}(1000/950)^{2/7} \tag{1.1}$$

$$\theta_B = 0.5(\,\theta_{850}+\theta_{700}) = 0.5[T_{850}(1000/850)^{2/7}+T_{700}(1000/700)^{2/7}] \tag{1.2}$$

$$\theta_C = \theta_{500} = T_{500}(1000/950)^{2/7} \tag{1.3}$$

$$r_A = r_{950} \tag{1.4}$$

$$r_B = 0.5(r_{850}+r_{700}) \tag{1.5}$$

$$r_C = r_{500} \tag{1.6}$$

Equivalent Potential Temperature Proxies (EPTP) are then calculated following one of the methods described by Davies-Jones (2009). These are called "proxies" because they are not true equivalent potential temperatures. A modification was made for simplification. It was noted that computing the temperature at the lifting condensation level was potentially complex and using instead 850 hPa temperatures produced negligible changes in final GDI values. Thus the formulas we used contain 850 hPa temperatures T_{850} [K] as follows:

$$EPTP_A = \theta_A e^{\left(\frac{L_o r_A}{C_{pd} T_{850}}\right)} \tag{1.7}$$

$$EPTP_B = \theta_B e^{\left(\frac{L_o r_B}{C_{pd} T_{850}}\right)} + \propto \tag{1.8}$$

$$EPTP_C = \theta_C e^{\left(\frac{L_o r_C}{C_{pd} T_{850}}\right)} + \propto \tag{1.9}$$

where $\propto = -10$ [K] is an empirical adjustment constant, $L_o = 2.69 \times 10^6$ [J kg^{-1}] is a latent heat constant adjusted for this formula and $c_{pd} = 1005.7$ [J kg^{-1} K^{-1}] is the specific heat of dry air at constant pressure.

A dimensionless EPTP Core Index (ECI) is then calculated. The ECI is an enhancement factor that addresses the potential for convection developing into the mid-troposphere, which is often sufficient for convection to progress into the upper tropical troposphere as the mid-lower troposphere is where the most stable layers often linger. A Mid-level EPTP factor (ME) and a Low-level EPTP factor (LE) are first calculated via (1.10) and (1.11):

$$ME = EPTP_C - \beta \tag{1.10}$$

$$LE = EPTP_A - \beta \tag{1.11}$$

where $\beta = 303$ [K] is an empirical constant that enhances EPTP variability. The ECI is then calculated via (1.12):

$$ECI = \left\{ \begin{array}{cc} \gamma \times LE \times ME & , LE > 0 \\ 0 & , LE \leq 0 \end{array} \right\} \tag{1.12}$$

where $\gamma = 6.5 \times 10^{-2}$ [K^{-1}] is an empirical scaling constant. The reasoning behind (1.12) is that if significant amounts of heat and moisture are available at mid-levels, and large values are also present at low-levels, the entire column is moist and the moisture is ground-based. This causes the potential for convection to increases non-linearly based on the departures of $EPTP_A$ and $EPTP_C$ from the β threshold.

A Mid-level Warming Index (MWI) is then calculated. The MWI accounts for the modulation of mid-level stability by the presence of a warm ridge (increase in stability) or a cool trough (decrease in stability). It is an inhibition index that only produces negative values as a function of how much warmer are 500 hPa temperatures from the $\tau = 263.15$ [K] threshold (~ -10 C). It can be calculated via (1.13):

$$MWI = \left\{ \begin{array}{cc} \mu \times (T_{500} - \tau) & , T_{500} - \tau > 0 \\ 0 & , T_{500} - \tau \leq 0 \end{array} \right\} \tag{1.13}$$

where $\mu = -7$ [K^{-1}] is an empirical scaling constant and T_{500} is the 500 hPa temperature in [K].

An Inversion Index (II) is then calculated as inversions play an important role in inhibiting tropical convection. The II is an inhibition factor as well (i.e. only negative values are allowed) and is calculated by considering two factors: stability across the inversion and dry air entrainment once convective cells penetrate the inversion layer. The Stability factor S is calculated via a simple difference of the temperatures at 950hPa (T_{950}) and 700 hPa (T_{700}) in [K]. The 950 hPa level was used since the inversion can sometimes be located at levels as low as 925 hPa, while other times rises up to 700-750 hPa. S is calculated via (1.14):

$$S = \sigma \times (T_{950} - T_{700}) \tag{1.14}$$

where $\sigma = 1.5$ [K^{-1}] is an empirical scaling constant. A Drying factor D is then calculated based on the difference of EPTP between layers A and B via (1.15). In the tropics, dry air associated with subsidence inversions has a strong signature in the EPTP field resulting in a sharp decrease of EPTP with height.

$$D = \sigma \times (EPTP_B - EPTP_A) \tag{1.15}$$

Both S and D are combined via (1.16) to calculate the II:

$$II = \left\{ \begin{array}{cc} 0 & , D + S > 0 \\ D + S & , D + S \leq 0 \end{array} \right\} \tag{1.16}$$

The GDI can then be calculated by simply adding the three indices calculated:

$$GDI = ECI + MWI + II \tag{1.17}$$

An optional correction is available to improve visualization around high mountain ranges. NOAA products online and Wingridds GDI equations use this correction. It is recommended when using GDI as a gridded forecasting tool, since it will tamper the excessive fictitious values that occur over tall mountains such as the Andes, Rockies, Mexican Sierras, Central American ranges or Hispaniola's ranges. This correction considers surface pressure P_{SFC} in [hPa]. Note that the GDI is not truly applicable over terrain located over 950 hPa due to the need of data at this level. Yet, validation has shown that GDI variability in some of these regions still leads to high correlations, some of which even exceed r^2=0.5. The surface pressure correction C_O is:

$$C_O = P_3 - \frac{P_2}{P_{SFC} - P_1} \tag{1.18}$$

where $P_1 = 500$ [hPa], $P_2 = 9000$ [hPa] and $P_3 = 18$ are all empirical constants tuned to improve GDI visualization over elevated terrain.

A version of the GDI optimized for visualization can be attained via (1.19):

$$GDIc = ECI + MWI + II + C_O = GDI + C_O \tag{1.19}$$